

1 The Time Sequence and Magnitude of Physical Effects of El Niño in the Eastern Pacific

E. FAHRBACH, F. TRILLMICH, and W. ARNTZ

1.1 Introduction

During the years 1982–84 different pinniped populations along the Pacific coasts of the Americas were heavily affected by climatic anomalies over a wide range of latitudes, extending from Alaska in the north to the south of Chile. These anomalies can be related to El Niño (EN), a climatological phenomenon characterized by anomalous conditions in the atmosphere and the ocean. It appears in a somewhat regular time sequence in the tropical Pacific, extends, however, less intensively to the mid-latitudes and beyond. During EN, anomalously warm water appears along the coast of Ecuador and Peru. As an example we show the deviation of the sea surface temperature from the long-term mean at Puerto Chicama (Fig. 1). The magnitude of this positive sea surface temperature (SST) anomaly can be used to classify the intensity of an EN.

Examination of colonial documents and shipboard records, collected since the Spanish exploration of South America, suggests that 47 strong to very strong EN events (with an SST anomaly over several months of 3 to 12 °C) have occurred during the last four and a half centuries. In addition, there were at least 32 moderate EN events with a temperature anomaly of 2 to 3 °C since 1800 (Quinn et al. 1987). Therefore, on average, an EN turns up once every 4 years, with strong to very strong events occurring once every 10 years, though the actual distribution shows significant deviations from true cyclicity (Fig. 2).

EN is related to the weather system over the entire Pacific Ocean, and in most cases, it coincides with a significant “Southern Oscillation”. The Southern Oscillation describes a seesaw-type change in the air pressure distribution over the South Pacific. It is quantified by an index which is based on the air pressure difference between Tahiti and Darwin, Australia. Under normal (non-EN) conditions, pressure is high over Tahiti and low over Darwin (Fig. 3).

A reversal of this pressure distribution, indicated by a negative Southern Oscillation Index, is usually followed by an EN event (Fig. 4). The causes of these changes in the pressure distribution lie in the instability of the large-scale atmospheric and oceanic circulation systems and their mutual interaction which we are just beginning to understand at present (Cane 1986; Enfield 1987; Graham and White 1988). Despite a dense network of meteorological and oceanographic stations measuring a multitude of parameters, it is still not possible to accurately predict EN events, even 2 months before they occur (Enfield and Wyrтки, pers. comm., Guayaquil Conference 1986). This fact became most evident when the secular EN of

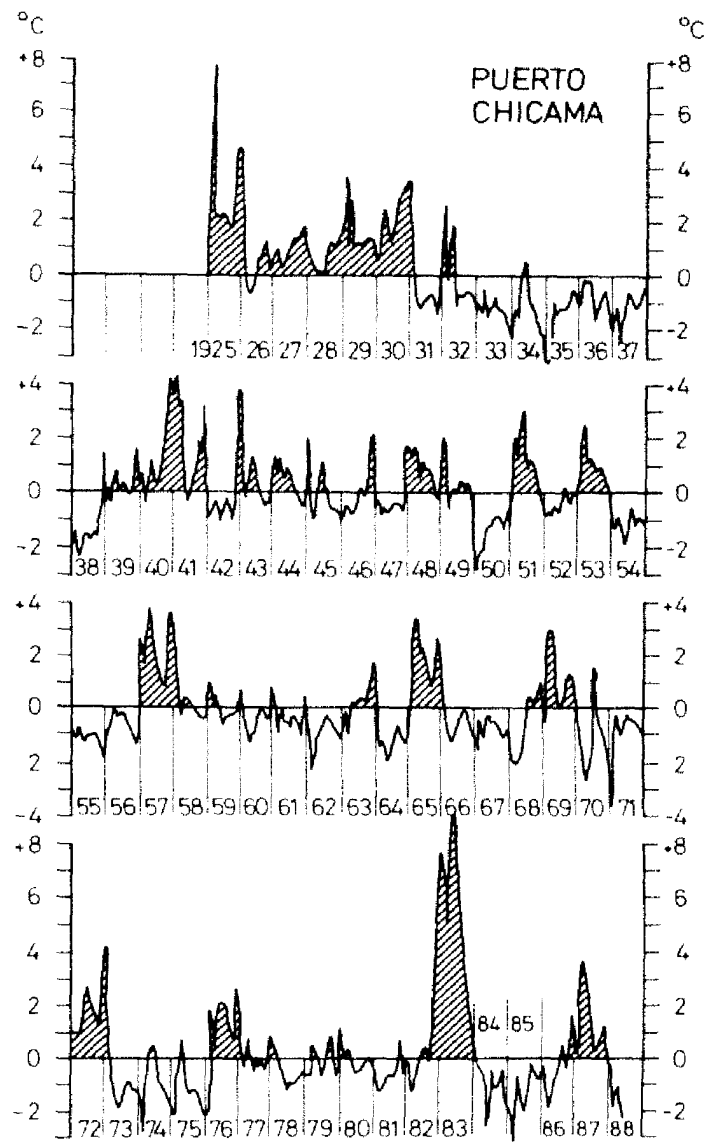


Fig. 1. Temperature anomalies (°C) at Puerto Chicama, Peru (7°42'S; 79°27'W) from 1925 to 1988 (Peruvian Navy, Callao). Positive anomalies are shaded

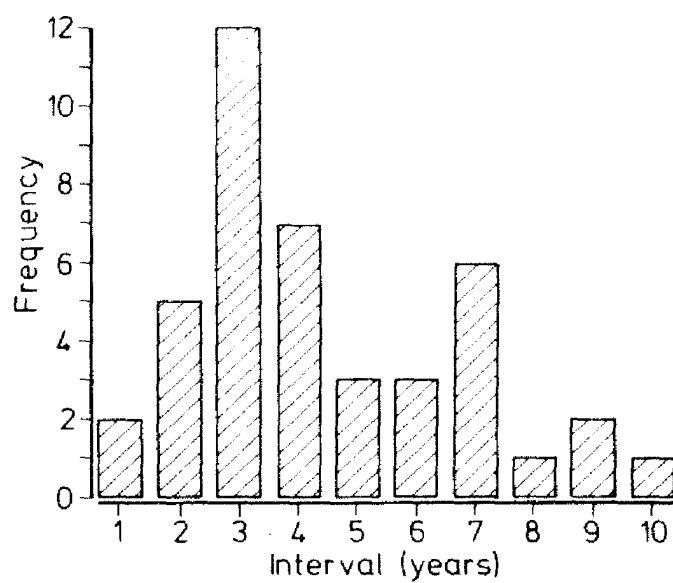


Fig. 2 Histogram of the intervals between EN events derived from data given by Quinn et al. (1987)

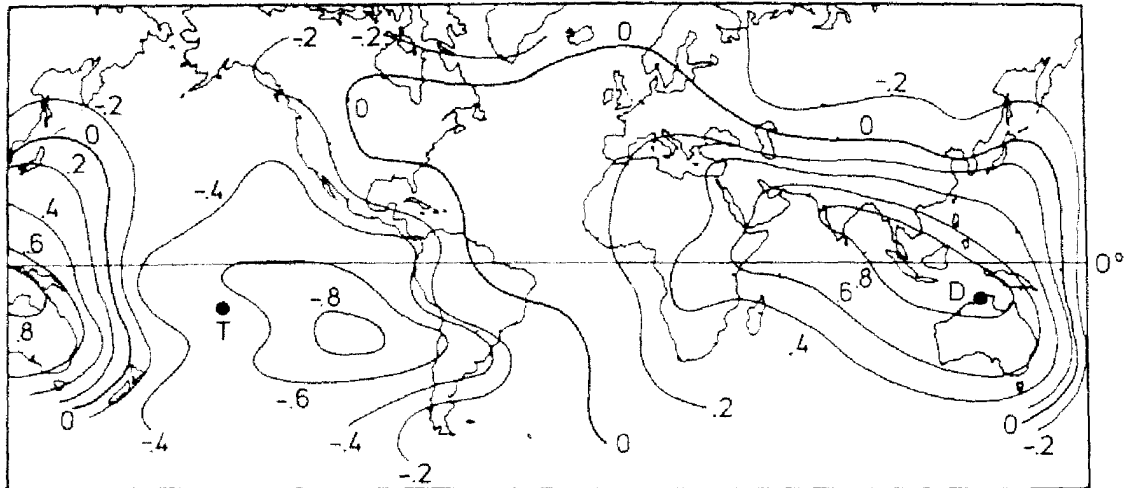


Fig. 3. The Southern Oscillation visualized by the local correlations of annual pressure anomalies with the ones at Djakarta, Indonesia, from Rasmusson (1984) after Berlage (1957). Tahiti (T) and Darwin (D) are highly correlated but with opposite sign

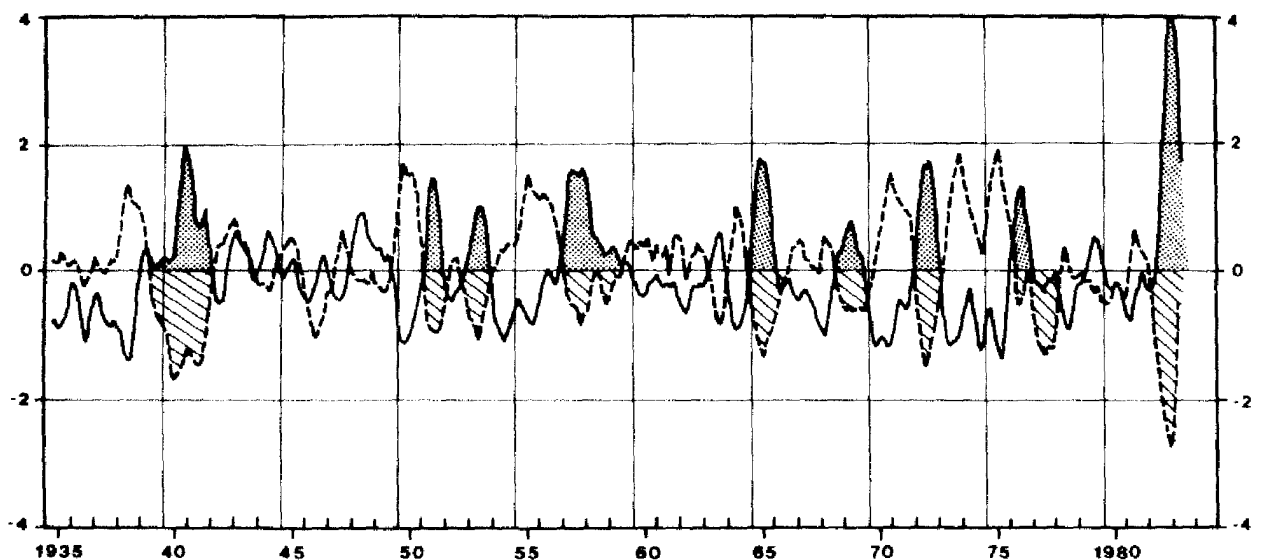


Fig. 4. Curves of the Southern Oscillation Index (SOI, *dashed line*) and sea surface temperature anomaly at Puerto Chicama (SST, *solid line*). Negative SOI coincides with positive SST anomaly (Rasmusson 1984)

1982–83 appeared completely without warning. This considerably dampened the high expectations of scientists who presumed the prediction problem had been solved through the work of Wyrski (1982; however, see Cane et al. 1986). Wyrski, analyzing wind and ocean conditions before and during the six EN events preceding that of 1982–83, established a number of rules which seemed to enable scientists to predict a coming event. While these rules are applicable for the six events before

1982–83 they failed in the case of EN 1982–83 due to a number of irregularities in the development of this event (Cane 1983, 1986; Rasmusson and Wallace 1983). At present, “hindcasting” is certainly more successful than forecasting; with the full amount of information at hand, the history of the more recent ENs, including the 1982–83 event, can be traced back to their origins.

We are not sure when EN-like events occurred for the first time off the American Pacific coast, and whether their appearance have since been continuous or interrupted. Evidence from sediment layers (Wells 1987), glaciers (Thompson and Mosley-Thompson 1986), and the accumulation of cadmium in corals (Shen and Boyle 1984) present proxy-indicators for EN-like events in the past. Based mainly on the study of fossil mollusks, deVries (1987) suspects that ENs may have existed for approximately 2 million years (since the Pliocene-Pleistocene boundary). An opposing view, i.e., a relatively recent “birth” of EN about 5000 years ago, has been suggested by archaeologists (Rollins et al. 1986). If deVries is correct, EN has been around for about 400 000 seal generations. At that time, the modern seal genera *Arctocephalus* and *Otaria* did not yet exist, but two Phocidae have been identified for the southern Peruvian coast (de Muizon 1981). However, according to present understanding a recent “birth” is more likely, because it is difficult to imagine that secondary variations like “El Niño” persist, if the climate is subject to dramatic changes like ice ages (Martinson, pers. comm.).

Direct and indirect changes brought about by EN events in the marine ecosystem greatly influence pinniped populations living in the eastern Pacific. This impact, both of an abiotic and biotic nature, occurred in a particularly severe and geographically widespread manner in the 1982–83 EN off South America (e.g. Barber and Chávez 1983; Cane 1983; Rasmusson and Wallace 1983; Philander 1983; Arntz 1984, 1986). Along the North American coast it was less dramatic but nevertheless the strongest event ever recorded.

First, we give a general description of some meteorological and oceanographic aspects of the EN phenomenon and then summarize information on changes of sea surface temperature, sea level, currents and temperatures at depth during the 1982–83 EN which are most likely to influence the distribution and abundance of organisms of importance to pinnipeds, either indirectly as food of their prey or directly as their own prey.

1.2 The Eastern Tropical Pacific Under Normal and EN Conditions

The eastern equatorial Pacific is unusually cold for tropical ocean areas. This is caused by equatorial and coastal upwelling and an influx of cold water with the Peru (Humboldt) Current (Fig. 5). The Peru Current is driven by SE trades which force surface waters away from the coast of Peru and Chile leading to local upwelling. This injects substantial amounts of cold water from deeper levels to the system. In contrast, the surface waters of the western Pacific are very warm (Fig. 6a). They are, as Cane (1986) states, the largest pool of very warm water in the world. It is maintained by the oceanwide trade winds, which drive currents westward (Fig. 5). In the tropics and subtropics the water is warmed on its way to the west due to the

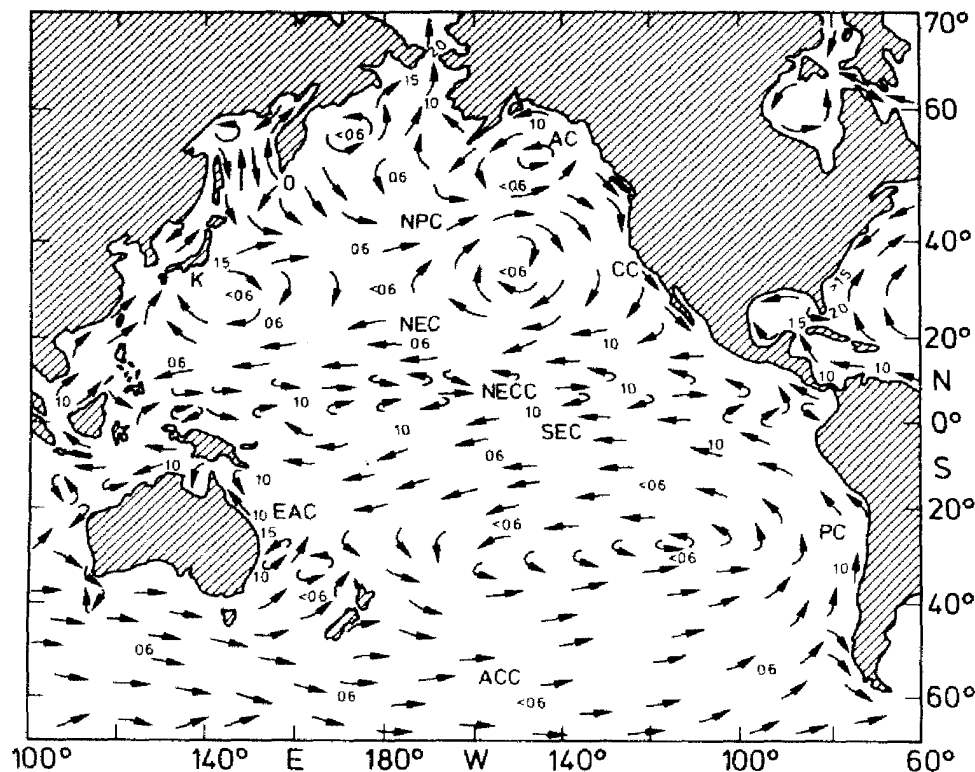


Fig. 5. Ocean Surface Currents of the Pacific Ocean in knots compiled from ship drift estimates for northern summer. (U.S. Navy 1977, 1979). Below the surface layer undercurrents can flow in the opposite direction; e.g. Peru Undercurrent and Equatorial Undercurrent. *AC* Alaska Current; *O* Oyashio; *NPC* North Pacific Current; *CC* California Current; *K* Kuroshio; *NEC* North Equatorial Current; *NECC* North Equatorial Counter Current; *SEC* South Equatorial Current; *EAC* East Australian Current; *PC* Peru (Humboldt) Current; *ACC* Antarctic Circumpolar Current

heat gain by solar radiation. Consequently, there is a significant gradient in SST along the equator.

By stirring the near-surface waters the winds generate a mixed layer. The entrainment of cold water from below induces a heat transport from the mixed layer to the deep ocean. If this heat loss is overcompensated by heat gain from the atmosphere above, the mixed layer warms up even if it deepens. Along the equator in the Pacific the mixed layer extends to about 100 m depth, becoming shallower to the east and almost disappears at the South American coast due to the penetration of cold water to the surface by upwelling. The lower boundary of the mixed layer is formed by a transition zone of increased vertical temperature gradients, the thermocline (Fig. 7a). In the tropical oceans the thermocline is comparably sharp.

The wind-induced water transport from east to west leads to an increase in sea level in the western Pacific which gives rise to a strong subsurface flow below the directly wind-driven layer from the west to the east (Fig. 7b). The Coriolis force concentrates this current on the equator. Therefore, it is generally called the Equatorial Undercurrent, however, in the Pacific, it is sometimes referred to as the Cromwell Current, after its discoverer. On its way along the equator it hits the Galápagos

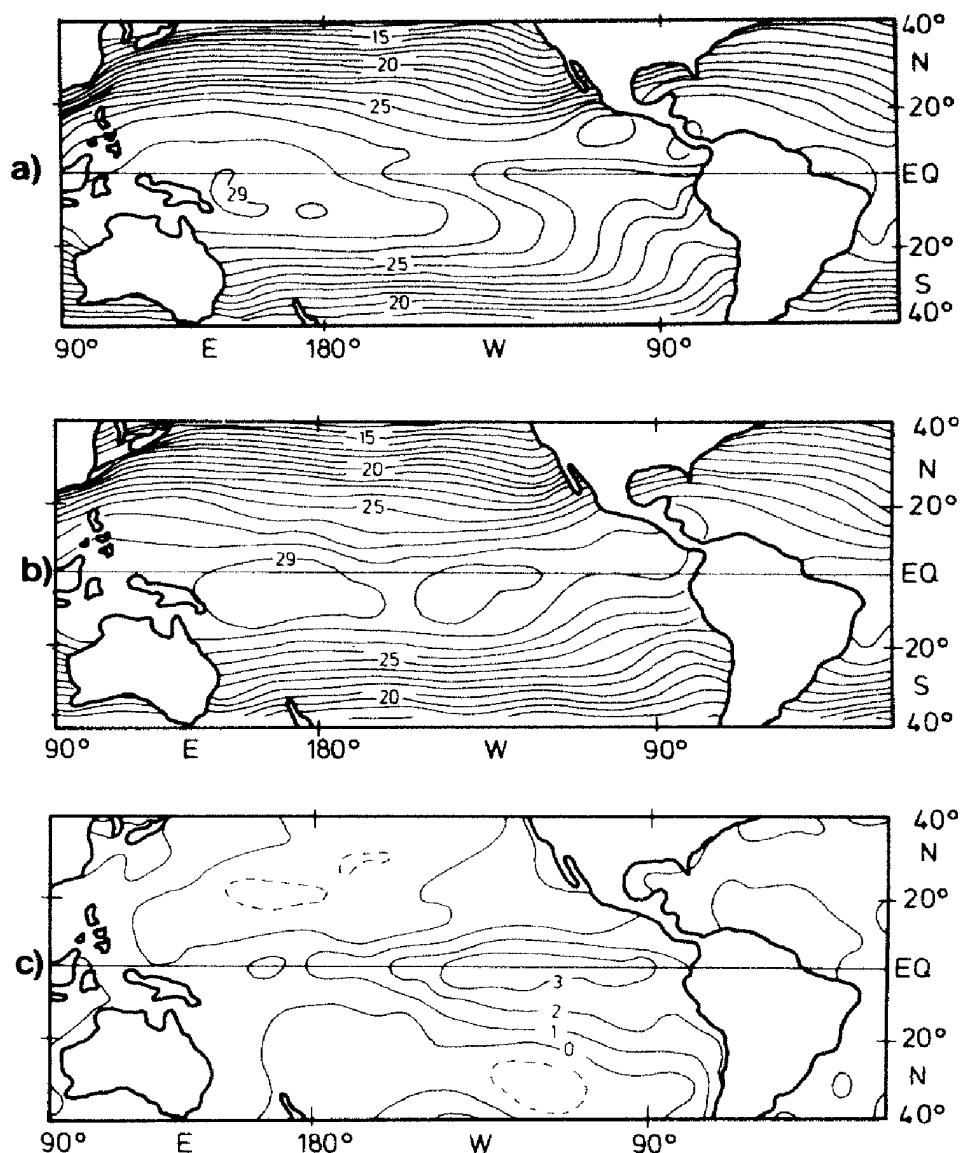


Fig. 6a-c. Sea surface temperature in °C in the tropical Pacific (after Rasmusson and Wallace 1983). **a** Climatological conditions from December to February. **b** Average conditions from December 1982 to February 1983. **c** Anomalies calculated as difference between **a** and **b**

Islands. Part of it surfaces and supplies large quantities of cold water to the upwelling ecosystem of the Galápagos. Its remainders are split into branches which surround the islands to the north and south (Houvenaghel 1984).

The SST distribution, with cold water in the east and warm water in the west, is associated with an atmospheric circulation along the equator: cold, relatively dry air flows from the eastern equatorial Pacific towards the west (Fig. 8). Over the warm water the air absorbs heat and moisture and rises. Some of this warm moist air returns to the east where it sinks, and some flows poleward at great altitudes. These patterns of air flow are related to high air pressure over the eastern tropical Pacific and lower pressure in the west.

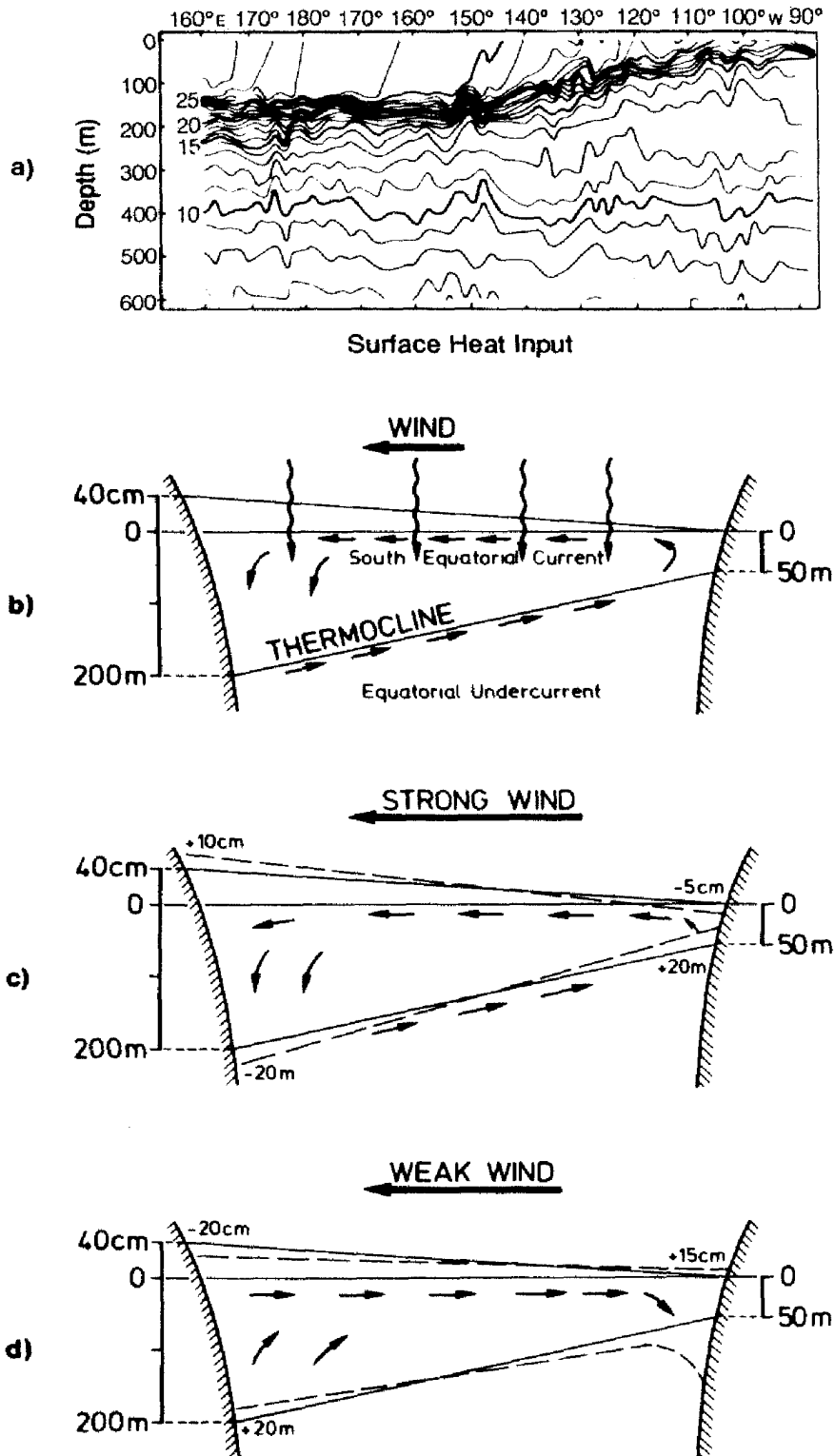


Fig. 7a-d. Thermal structure of the equatorial Pacific (after Colin et al. 1971; Wyrski 1982). **a** Observed structure, schematic representations; **b** normal conditions; **c** strong trade winds pile up water in the west; **d** after relaxation of the wind water sloshes to the east inducing sea level rise and thermocline deepening

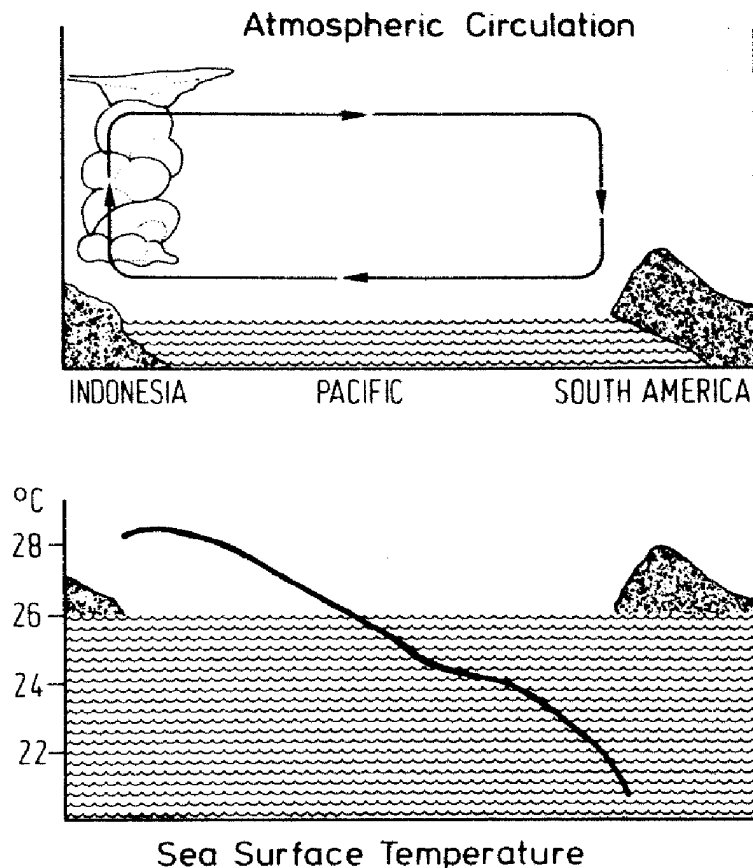


Fig. 8. Schematic representation of the Walker circulation in the atmosphere (*above*) and sea surface temperature along the equator (*below*). Warm moist air is rising over the warm western Pacific. In the east cool air is sinking (after Wyrtki 1982)

When the trade winds weaken, which is a normal event at the beginning of each southern summer, the cool Peru Current slows down and the upwelling of cold water is diminished. However, the surface heating rate is maintained, so the surface waters of the equatorial Pacific are warmed. The increasing volume of warm water depresses the thermocline from about 30 m during the cold season to roughly 50 m during the warm season. In normal years though, little warming is noticeable along the coast of Peru below 50 m depth (Chávez 1987). This is the normal situation in January (austral summer). The warm period ends towards April when the SE winds gain strength and cool conditions return. Even during this cool period occasional surface and subsurface warming occurs between April and July which may last for about 1 month (Chávez 1987).

During an EN event, instead of cooling, warming continues. This is caused by backwashing of warm waters which were accumulated in the western Pacific by anomalously strong trades during the prior year (Fig. 7). The relaxation of the trades excites wave patterns trapped along the equator which are called Kelvin waves. Due to these Kelvin waves sea level rises 10 to 30 cm and the thermocline is depressed several tens of meters. The surface current to the west is consequently reduced or

even reversed. As Kelvin waves propagate with a velocity of 2 to 3 m/s they reach the South American coast 6 to 8 weeks after their excitation in the western Pacific (Fonseca 1985). When they arrive at the coast they deviate and propagate along the coast to the south and north, generating sea level rise and thermocline depression well before atmospheric anomalies can be observed. The observation that relaxing trade winds in the western Pacific induce sea level rise on the eastern coast gave birth to the notion of "teleconnection". This relationship can now be explained by propagation of Kelvin waves.

In the atmosphere anomalies occurring in the tropical Pacific excite perturbations of the extratropical weather systems by wave propagation. The atmospheric "teleconnections" are due to so-called planetary waves.

The depression of the thermocline during an EN is of considerable importance to the coastal upwelling ecosystem. The upwelled water originates from depths of 50 to 100 m which are normally below the thermocline and therefore cold and rich in nutrients. However, during EN, due to the deepening of the thermocline, upwelled water originates from the surface mixed layer and is warm and poor in nutrients. Consequently, even if the trade winds do not slacken along the coast and upwelling continues, the nutrient supply necessary to enhance primary production is interrupted (Barber and Chávez 1986).

Prior to the 1982–83 EN, several EN events were used to describe the so-called *canonical* (composite) EN (Wyrski 1982). The 1982–83 and before it the 1940–41 ENs show a number of peculiarities which do not fit the previously established rules. Prior to the 1982–83 EN the trade wind failed to blow stronger than normal which is observed during the canonical EN and leads to the accumulation of warm water in the west (Kerr 1983). Consequently, no sea level elevation was observed there (Cane 1983). Warming on the equator started in the central Pacific and spread eastward to the South American coast instead of starting there. The 1940–41 EN was quite similar (Rasmusson and Wallace 1983), although at that time the temperature anomalies were much weaker than during the 1982–83 event. Furthermore, the 1982–83 EN lasted to mid-July 1983 instead of ending in April as EN events off South America usually do. Off North America, unusual warming even continued into 1984 (Wooster and Fluharty 1985).

1.3 Abiotic Effects of EN 1982–83 in the Eastern Tropical and Southern Pacific

In the eastern tropical and southern Pacific the effects of the 1982–83 EN were most pronounced. The mean sea surface temperature from December 1972 to February 1983 is shown in Fig. 6b, the anomaly from the climatological conditions in Fig. 6c. We attempt here to give a brief description of the sequence and extent of the effects from the Galápagos to Chile, the areas for which we have data on pinnipeds. Effects along the North American coast are presented in the next section.

1.3.1 Galápagos and Mainland Ecuador

The monthly mean sea level in Galápagos began to rise above the long-term average in October 1982, reached a first maximum of 37 cm above normal in December and a second one of 33 cm in May 1983. The absolute maximum value was 47 cm above normal on 6 January 1983 (Wyrski 1985). Along the equator west and east of the Galápagos the usual westward flow stopped and the surface current to the west of the Galápagos was reversed from about mid-September to mid-December 1982 carrying warm water from the west into the Galápagos area. The Equatorial Undercurrent seemed to disappear or even partly reverse during the same period (Firing et al. 1983; Taft 1985; Jiménez and Intrigo 1986) but returned again in January 1983. The 15 °C isotherm normally found between 50 and 100 m was deeper than 200 m in December 1982. Water warmer than 25 °C extended to 100 m and surface temperatures rose to 27 °C. In March 1983 the 15 °C isotherm ascended to about 150 m depth and the 25 °C isotherm was at 50 m. The surface water, however, continued warming and reached more than 30 °C at times. In late June 1983 the waters west of the Galápagos Archipelago cooled rapidly while the northern and eastern islands (Wolf, Darwin and San Cristobal) were bathed by warm waters for one more month. Apparently, the cool waters took much longer to reach north of the equatorial front where these three islands are situated (Hayes 1985). In October 1983 all these parameters had returned to normal, while sea level and SST were even lower than the long-term mean for Galápagos. Off mainland Ecuador, sea level was more than 10 cm above the 1975–1981 average between May and September 1982, rose sharply from October and reached a first peak of 30 cm in January 1983. After a decline in the following months, a second peak of about the same height was reached in May 1983 before a steady decline which culminated in less than 10 cm below average values in December 1983 (Cucalón 1987).

During the first peak of the event, in February 1983, SSTs off Ecuador were 4 °C higher than in 1981 which is considered a normal year. Interestingly, temperatures at 50 m were 9 °C higher than 1981, i.e., the subsurface signal of EN in this area was clearly stronger than the surface signal. The 25 °C isotherm which was found at the surface in February 1981 sank to 50 m in February 1983; at the same time, the 16 °C isotherm declined by 80 to 130 m. In previous ENs, a similar but weaker deepening of the 16 °C isotherm had been observed; in February 1972 it was at 100 m, in February 1976 at 80 to 100 m depth. By April to May 1983, SST off Ecuador continued to increase; the highest SST anomalies (up to 6.3 °C, which corresponds to a temperature of 29.5 °C) were observed from May to July 1983. Thereafter, the system gradually recovered (Cucalón 1987).

1.3.2 Peru and Chile

This description of the event largely follows Arntz (1984, 1986), Fonseca (1985), and Guillén et al. (1985). The front of the Kelvin wave packet reached the South American coast, at 10°S, on 7 October 1982, about 6 days after passing the Galápagos. A sea level rise of up to 40 cm in Ecuador and Peru was similar to that

observed in Galápagos. Maxima of around 30 cm were reported as far south as Antofagasta (23°S) in Chile (Fonseca 1985). The sea level along roughly 25° of latitude peaked in January 1982 and again between April and June 1983. The southern extent of the rise in sea level is not adequately documented.

The sea level elevation led to an enhancement of poleward flow and therefore advection of warm water. The Peru Current was strongly reduced and even reversed its direction at times during April and May 1983. At 10 °S, the Peru Undercurrent increased its poleward speed from 4.2 to 25.3 cm/s (the 64-day average) with peak velocities of 35.8 cm/s. After 10 July 1983 this current returned again to its former slower speeds (R.L. Smith 1983, 1984).

In November 1982 SSTs reached 23 to 26 °C along the Peruvian coast, i.e., were 4 to 6 °C warmer than normal. The warm waters extended as far south as Arica in Chile (18°30'S). Larger cold water areas were left only off San Juan (15°20'S) and Atico (16°10'S) with SSTs less than 20 and 19 °C, respectively. During December these cooler upwelling areas shrunk drastically. Efficient upwelling resulting in nutrient supply to the near-surface waters was restricted to a few areas close to the coast. Further offshore upwelling did not reach below the thermocline. Consequently, the upwelled water was poor in nutrients. SSTs reached their first maximum in January 1983, with values predominantly between 26 and 30 °C north of Callao. In the south of Peru this peak was delayed until March attaining 23 to 27 °C. In northern Chile SST exceeding, from January to March, the long-term mean by 4.5 °C, increased to 28 °C. The warming extended southward to at least 37 °S (Kelly 1985). By June temperatures had returned to near normal in northern Chile and southern Peru where they achieved the long-term mean again in September (Fuenzalida 1985). North of 14 °S, however, warming continued until June, with a second peak during April and May 1983. After July 1983, the SST dropped drastically but had gained normal levels again by October 1983, although slight positive anomalies persisted locally north of Callao. This anomalous SST was accompanied by large-scale warming at depth. The 15 °C isotherm, normally at about 50 m depth nearshore, sank to more than 200 m in December 1982, but returned to about 130 m in February 1983 (Guillén et al. 1985). In northern Chile it dropped to 150 m depth, off Antofagasta to about 100 m (Blanco and Díaz 1985). At the peak of the event, oceanographers traced the warming down to 800 m (Guillén et al. 1985).

Dissolved oxygen was slightly reduced in the surface layers whereas at depth the warm intruding water induced an oxygen increase. A comparison of the "Humboldt" cruises 8103/04 (normal conditions) and 8212/8301 (EN) indicated a three- to seven fold increase of dissolved oxygen at the seafloor off northern and central Peru during EN (Arntz et al. 1985). An unusually well-oxygenated layer with values up to 2.5 ml/l O₂ was thus established to a depth of 150 to 200 m (Guillén et al. 1985). In contrast to the conditions off Peru, during the later phase of EN, the surface waters off northern Chile were poor in oxygen with values less than 1 ml/l (Alvial 1985; Fonseca 1985; Fuenzalida 1985).

Simultaneous with the warming of the sea, the well-known effect of a much increased rainfall in a normally arid coastal region, with both catastrophic and positive consequences, was observed.

1.4 Effects of EN Along the Coast of North America

Important populations of pinnipeds live along the coast of North America in the California and Alaska Currents. The North Pacific Current system is mainly wind-driven. Hence, its flow and position relative to the coast strongly depends upon the position of the Aleutian low and the Pacific high.

Under normal conditions, the California Current (0–200 m deep) carries water southeastward parallel to the coast throughout the year (Fig. 5). Its water comes from the Subarctic Pacific and the West Wind Drift and enters the current at about 48 °N. This water is cool, of low salinity and high in oxygen and phosphate. Volume transport is maximum in summer and minimum in winter and spring. Upwelling intensity along the coast parallels these changes in the intensity of current flow: most upwelling takes place during the spring and summer months. From October through January, the California Current is displaced about 100–150 km offshore. During this time, the California Countercurrent (which further north merges with the Davidson Current) transports warmer water nearshore northward.

Effects of the tropical EN in 1982–83 were transmitted to the coast of North and Central America through two processes, the so-called teleconnections.

1. Poleward propagation of Kelvin waves resulting in an elevation of coastal sea level, an increase in SST and stronger poleward flow.
2. Planetary waves in the atmosphere induce variations in the wind fields due to changes in the pressure systems over the central and north Pacific, in particular, intensification of the Aleutian low. They lead to the relaxation of coastal upwelling (Hamilton and Emery 1985; Norton et al. 1985).

The relative role of these two processes in the anomalies observed along the coast of North America is still not completely understood.

The sequence of the EN effects along the coast of North America varied with latitude. This was more noticeable with respect to the end of EN effects than to their onset. An increase of nearshore (0 to 200 km) SSTs was noted from Panama to Canada in October 1982 (Wooster and Fluharty 1985). In the northern Gulf of Alaska the first signs were delayed until January 1983 (Royer 1985). Peak anomalies generally occurred in the period of March to May 1983 at all latitudes from Mexico to Canada but were retarded in the Gulf of Alaska until June and July 1984 in surface waters. Warming observed in the Gulf of Alaska in 1983 and 1984 was part of a long-term warming trend and may not be an EN signal (Royer, unpubl.).

The width of the affected coastal strip, depth range and strength of the temperature signal also varied strongly with latitude. The effect of a warming event on the biological system is different from area to area. Whereas biological productivity is reduced in the California Current by the warming, it is enhanced in the Gulf of Alaska and Bering Sea (McLain 1983). In reviewing the physical effects we will proceed from the south to the north. Most of the data mentioned in the following stem from the book edited by Wooster and Fluharty (1985), where additional useful information can be found.

Off Mexico, a first maximum in sea level elevation occurred in December 1982 and a second one from March to May 1983 (Galindo et al. 1986). Warm water appeared even earlier in July 1982 near Guerrero (16° to 18 °N) (Ramos et al. 1986). SST obtained its first maximum from October to December 1982 and a second one from March to May 1983 (Galindo et al. 1986). The warm water layer extended to depths of more than 100 m (Gallegos et al. 1986). During EN more subsurface water entered the Gulf apparently because of the strengthened Costa Rica Current. This induced intense mixing with waters from the inner Gulf making more nutrients available for phytoplankton growth (Baumgartner et al. 1987).

In San Diego (southern California) oceanographers first began to notice the rise in sea level in June and July 1982 (Simpson 1984b). Along the western coasts of Baja and Alta California SSTs climbed above average in fall 1982. The rise in SST extended from the coast to about 500 to 1000 km offshore (Norton et al. 1985). Effects were strongest in winter 1982–83 and again in winter 1983–84. Further offshore (more than 1000 km) SSTs were lower than normal in winter 1982–83, but near normal in 1983–84. The warm anomalies evidently affected not only the surface but extended to a depth of more than 200 m. At 100 m depth a dramatic warming was observed in winter 1982–83. After cooling in spring and summer 1983 a second major warming occurred in late summer, fall and winter of 1983. Between 29° and 34 °N positive temperature anomalies at 100 m depth varied from 2.3 to 3 °C in 1982–83 and achieved about 1.8 °C in the following fall returning to near normal in winter 1983–84.

During winter and spring of 1983, the storm tracks associated with the prevailing westerlies moved southward because of the expanded Aleutian low and caused increased rainfall and severe floods in most of California. Extreme southerly coastal winds associated with the intensified Aleutian low suppressed coastal upwelling and caused much damage through increased swell in combination with the rise in sea level of more than 30 cm. In February and March 1983 sea level showed a monthly anomaly of 26 cm at San Francisco and about 15 cm in southern California. The northward coastal countercurrent flow increased and brought unusually warm water far north. Simultaneously, the California Current transported less subarctic water southward. These combined factors produced the strong warming event along the coast of California (Norton et al. 1985).

Further north, along the coast of Oregon, Washington and British Columbia a coastal band of warm water of about 30 km width was established in November 1982 and increased to a width of about 300 km in February and March 1983. Concurrently, the monthly average sea level rose about 30 cm above the long-term mean. SST anomalies near the coast exceeded 3 °C. The anomalies peaked around April 1983 and SSTs warmer than normal were maintained throughout summer. Most of the surface warming had disappeared by August 1983. It vanished completely along the coast of Washington and Oregon in December 1983, but continued into 1984 further north. Off British Columbia subsurface water warmed down to 800 m depth with a maximum anomaly at 200 m. Warming of up to 5 °C between 50 and 100 m depth was observed over the shelf in November 1982 and extended for about 200 km offshore in March 1983. By the end of June 1983 the subsurface layers had returned to near-normal temperatures, but along the Oregon coast anomalies were

still recorded in July 1983 affecting a depth of more than 500 m (Tabata 1984, 1985; Huyer and Smith 1985).

Data from the northern Gulf of Alaska (Royer 1985) indicate that the event lasted over 1 year in Alaskan waters and invoked temperature anomalies greater than 2 °C at a depth of 100 to 200 m and of 5 °C at the surface. Warming began in early 1983 and anomalies lasted at least until June 1984. Positive temperature anomalies occurred at all depths over the shelf and had decreased again to less than 1 °C by June 1984. The Alaskan warming was clearly related to the intensification of the Aleutian low and its movement to the southeast, resulting in the flow of warmer waters from the North Pacific into the northern Gulf of Alaska. Sea level height in the Gulf increased by more than 20 cm from January to March 1983 (Norton et al. 1985) reflecting the increase in temperature of the adjacent water mass.

Warming of the Bering Sea began in November 1982 when an anomaly of 2 °C extended from the north Pacific Ocean to the Kamchatka peninsula, and northward toward the Bering Strait (NOAA 1982–84). In December, the anomaly extended eastward across the entire Bering Sea. The major warming of 3 °C in the southern Bering Sea lasted from January through April 1983, followed by a secondary peak not exceeding 1.5 °C and lasting from July 1983 through March 1984. Thereafter, it decreased below 1 °C. In comparison to the Gulf of Alaska, the warming of the Bering Sea began and ended slightly earlier, was slightly shorter, and its peak anomaly was about 2 °C less. The warming mechanism of the Bering Sea was probably an increased southerly air flow from the north Pacific Ocean associated with changes in the strength and location of the Aleutian low (Niebauer 1985). In later papers Niebauer (1988) and Niebauer and Day (1989) stated that the 1982–83 EN had little apparent effect in the Bering Sea. This is explained by the extreme eastern position of the Aleutian low and northerly winds during this period.

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